

WAVES

- Wave height (H) refers to the overall vertical change in height between the wave crest (or peak) and the wave trough. The wave height is twice the wave amplitude (a). Wavelength (L) is the distance between two successive peaks (or two successive troughs). Steepness is defined as wave height divided by wavelength (H/L)

Characteristics of wave motion

- A wave transfers a disturbance from one part of a material to another. (The disturbance caused by dropping a stone into a pond is transmitted across the pond by ripples.)
- The disturbance is propagated through the material without any substantial overall motion of the material itself.
- The disturbance is propagated without any significant distortion of the wave form. (A ripple shows very little change in shape as it travels across a pond.)
- The disturbance appears to be propagated with constant speed.

What is transported?

The answer, 'energy', provides a reasonable working definition of wave motion- a means whereby energy is transported across or through a material without any significant overall transport of the material itself.

TYPES OF WAVES

All waves can be regarded as progressive waves, in that energy is moving through, or across the surface of, the material. The so-called standing wave, of which the plucked guitar string is an example, can be considered as the sum of two progressive waves of equal dimensions, but travelling in opposite directions.

Waves which travel through the material are called body waves. Examples of body waves are sound waves and seismic P- and S-waves, but our main concern in this Volume is with surface waves (surface waves are those which occur at the interface between atmosphere and ocean, caused by the wind blowing over the sea)

In the case of surface waves on water, there are two such restoring forces which maintain wave motion:

- 1 The gravitational force exerted by the Earth.
- 2 Surface tension, which is the tendency of water molecules to stick together and present the smallest possible surface to the air. (capillary waves, wave length really small)

At the sea-surface, most of the transferred energy results in waves, although a small proportion is manifest as wind-driven currents. Waves obtain energy from the wind by virtue of pressure differences caused by the sheltering effect provided by wave crests. To work it requires to:

- 1 Wind speed exceeds wave speed.
- 2 Wind speed exceeds 1 m s^{-1} .
- 3 The waves are steep enough to provide a sheltering effect.

The sheltering effect is at a maximum when wind speed is approximately three times the wave speed. In general, the greater the amount by which wind speed exceeds wave speed, the steeper the wave.

- Wind stress exerted upon surface
- Wave growth due to pressure differences at either side of the crest
- Waves cease to grow because of energy loss due to current generation, friction (heat/sound), white-capping

Breaking reasons

- 1 Some of the wind energy is transferred to the ocean surface via a tangential force, producing a surface current.
- 2 Some wind energy is dissipated by friction, and is converted to heat and sound.
- 3 Energy is lost from larger waves as a result of white-capping, i.e. breaking of the tip of the wave

crest because it is being driven forward by the wind faster than the wave itself is travelling.

FULLY DEVELOPED SEA

Provided the fetch is extensive enough and the wind blows at constant speed for long enough, an equilibrium is eventually reached, in which energy is being dissipated by the waves at the same rate as the waves receive energy from the wind. Such an equilibrium results in a sea state called a fully developed sea, in which the size and characteristics of the waves are not changing.

WAVE HEIGHT AND WAVE STEEPNESS

Many oceanographers use the significant wave height, $H_{1/3}$, which is the average height of the highest one-third of all waves occurring in a particular time period. In any wave record, there will also be a maximum wave height, H_{max}

- The absolute height of a wave is less important to sailors than is its steepness (H/L). As mentioned in Section 1.1.2, most wind-generated waves have a steepness in the order of 0.03 to 0.06. Waves steeper than this can present problems for shipping, but fortunately it is very rare for wave steepness to exceed 0.1. In general, wave steepness diminishes with increasing wavelength.*

SURFACE WAVE THEORY

It is important to realize that the orbits are only approximately circular. There is a small net component of forward motion, particularly in waves of large amplitude, so that the orbits are not quite closed, and the water, whilst in the crests, moves slightly further forward than it moves backward whilst in the troughs. This small net forward displacement of water in the direction of wave travel is termed wave drift (see Figure 1.8(b)). In shallow water, where depth is less than half the wavelength and the waves 'feel' the sea-bed, the orbits become progressively flattened with depth (Figure 1.8(c) and (d)).

Alternating bands of smooth and rippled surface water at intervals of a few hundred metres may thus indicate the passage of an internal wave, but whether ripples represent convergences or divergences depends upon local conditions.

ASSUMPTIONS IN SURFACE WAVE THEORY

- 1 The wave shapes are sinusoidal.*
- 2 The wave amplitudes are very small compared with wavelengths and depths.*
- 3 Viscosity and surface tension can be ignored.*
- 4 The Coriolis force (see Section 2.3) and vorticity (Section 1.1.1), which result from the Earth's rotation, can be ignored.*
- 5 The depth is uniform, and the bottom has no bumps or hummocks.*
- 6 The waves are not constrained or deflected by land masses, or by any other obstruction.*
- 7 That real three-dimensional waves behave in a way that is analogous to a two-dimensional model.*

WAVES DISPERSION AND GROUP SPEED

This separation of waves by virtue of their differing rates of travel is known as dispersion, and Equation 1.3 ($c = \sqrt{gL / 2\pi}$) is sometimes known as the dispersion equation, because it shows that waves of longer wavelength (L) travel faster than shorter wavelength waves. The speed of the band, called the group speed, is about half the wave speed of the individual ripples which travel through that band.

As the water becomes shallower, wavelength becomes less important, and depth more important, in determining wave speed. As a result, in shoaling water, wave (phase) speed decreases, becoming closer to group speed. Eventually, at depths less than $L/20$, all waves travel at the same depth-determined speed, there will be no wave-wave interference, and therefore in effect each wave will

represent its own 'group'. Thus, in shallow water, group speed can be regarded as equal to wave (phase) speed.

WAVE ENERGY

The energy possessed by a wave is in two forms:

- 1 kinetic energy, which is the energy inherent in the orbital motion of the water particles; and
- 2 potential energy possessed by the particles as a result of being displaced from their mean (equilibrium) position.

PROPAGATION

It follows that the energy is contained within the wave group, and travels at the group speed. The rate at which energy is supplied at a particular location (e.g. a beach) is called wave power, and is the product of group speed (C_g) and wave energy per unit area (E), expressed per unit length of wave crest.

ATTENUATION

Wave attenuation involves loss or dissipation of wave energy, resulting in a reduction of wave height. Energy is dissipated in four main ways:

- 1 White-capping, which involves transfer of wave energy to the kinetic energy of moving water, thus reinforcing the wind-driven surface current (Section 1.1.2).
- 2 Viscous attenuation, which is only important for very high frequency ~capillary waves, and involves dissipation of energy into heat by friction between water molecules.
- 3 Air resistance, which applies to large steep waves soon after they leave the area in which they were generated and enter regions of calm or contrary winds.
- 4 Non-linear wave-wave interaction, which is more complicated than the simple (linear) combination of frequencies to produce wave groups as outlined.

SWELL

Such long waves are known as swell, which is simply defined as waves that have been generated elsewhere and have travelled far from their place of origin. If you look out to sea on a calm day, the waves that you see will be swell waves from a distant storm.

The waves with the longest periods travel fastest, and progressively out-distance waves of higher frequencies (shorter periods). Near to the storm, dispersion is likely to be minimal (Figure 1.11(b)), but the further one moves from the storm location, the more clearly separated waves of differing frequencies become, resulting in the regular wave motions we know as swell.

WAVES APPROACHING THE COAST

SHOALING

However, it is essential to realize that while the energy and height of individual waves will increase as they enter shallow water, the rate of supply of wave energy (wave power, Section 1.4.1) must remain constant (ignoring frictional losses). As mentioned earlier, when waves move into shallow water, the waves begin to 'feel the bottom', the circular orbits of the water particles become flattened (Figure 1.8(c) and (d)), and wave energy will be dissipated by friction at the sea-bed, resulting into-and-fro movement of sediments. The gentler the slope of the immediate offshore region, the sooner the incoming waves will 'feel' the bottom, the greater will be the friction with the sea-bed and the greater the energy loss before the waves finally break.

REFRACTION

Refraction of waves in progressively shallowing (shoaling) water can be described by a relationship equivalent to Snell's law, which describes refraction of light rays through materials of different refractive indices. Rays can be drawn perpendicular to the wave crests, and will indicate the direction of wave movement. The angle between these wave rays and lines drawn perpendicular to

the depth contours can be related to wave speeds at various depths.

WAVE BREAKING MECHANISMS

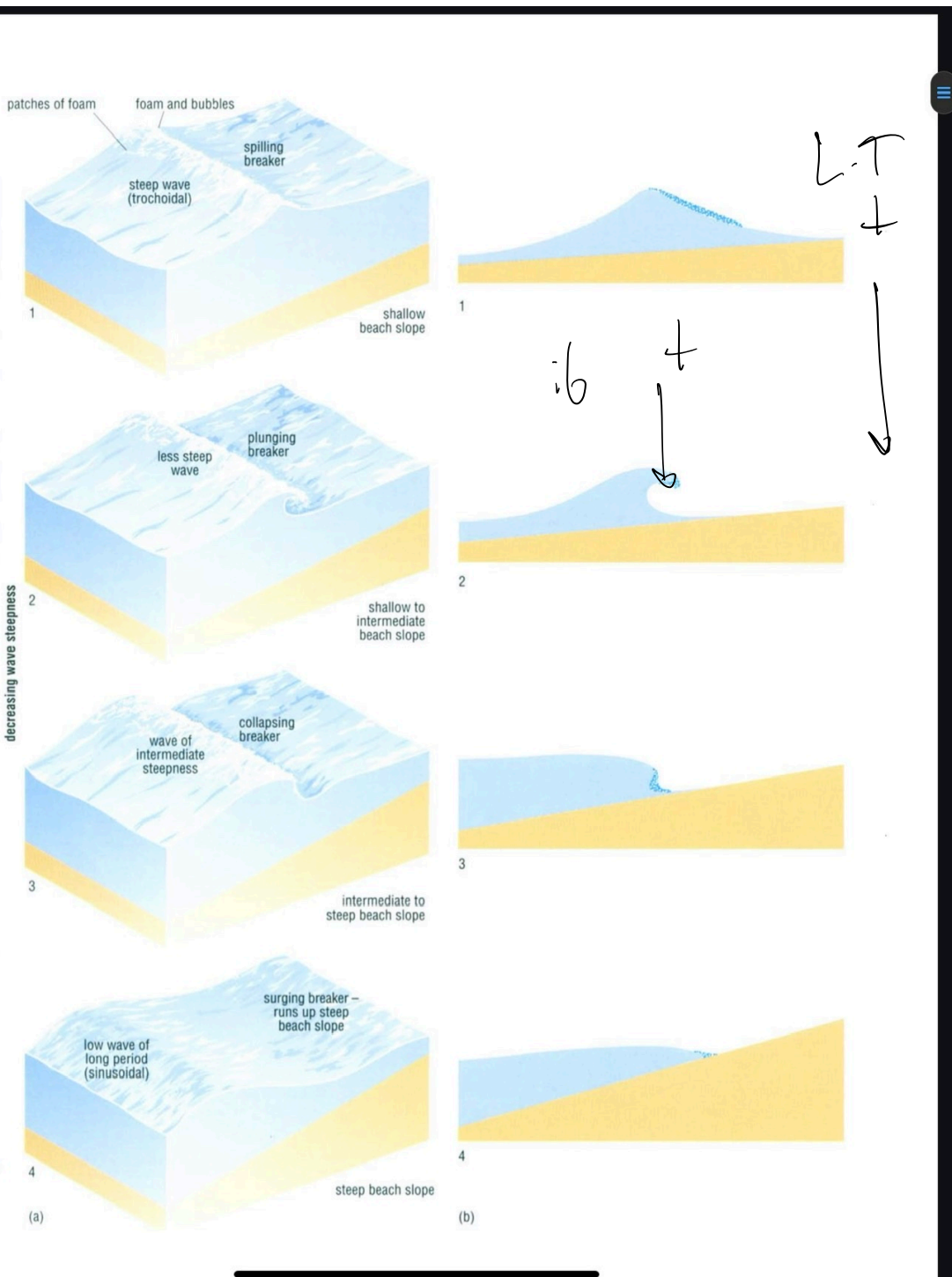
1 Spilling breakers are characterized by foam and turbulence at the wave crest. Spilling usually starts some distance from shore and is caused when a layer of water at the crest moves forward faster than the wave as a whole.

2 Plunging breakers are the most spectacular type. The classical form, much beloved by surf-riders, is arched, with a convex back and concave front. The crest curls over and plunges downwards with considerable force, dissipating its energy over a short distance. Plunging breakers on beaches of relatively gentle slope are usually associated with the long swells generated by distant storms

3 Collapsing breakers are similar to plunging breakers, except that the waves may be less steep and instead of the crest curling over, the front face collapses. Such breakers occur on beaches with moderately steep slopes, and under moderate wind conditions, and represent a transition: from plunging to surging breakers.

plun~in~to sur~in~ breakers.

4 Surging breakers are found on the very steepest beaches. Surging breakers are typically formed from long, low waves, and the front faces and crests thus remain relatively unbroken as the waves slide up the beach.



The way breaker shape changes from top to bottom of the picture depends upon:

- 1 Increasing beach slope (if considered independently from wave characteristics).
- 2 Increasing wavelength and period and correspondingly decreasing wave steepness, if these characteristics are considered independently of beach slope.

• SEICHE

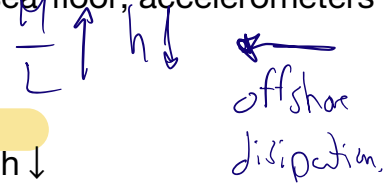
A seiche is a standing wave, which can be considered as the sum of two progressive waves travelling in opposite directions (Section 1.1).

Seiches :an occur in lakes, and also in bays, estuaries or harbours which are open to -he sea at one end.

SHALLOW WATER

Waves in shallow water may be refracted. Variations in depth cause variations in speed of different parts of the wave crest; the resulting refraction causes wave crests to become increasingly parallel with bottom contours. The energy of refracted waves is conserved, so converging waves tend to increase, and diverging waves to diminish, in height. Waves in shallow water dissipate energy by frictional interaction with the sea-bed, and by breaking. In general, the steeper the wave and the shallower the beach, the further offshore dissipation begins. Breakers form a continuous series from steep spilling types to long-period surging breakers.

Waves are measured by a variety of methods, e.g. pressure gauges on the sea floor, accelerometers in buoys on the sea-surface, and via remote-sensing from satellites.



FROM SLIDES

- Wave height changes when waves enter shallow water • Generally: $H \uparrow$ as $h \downarrow$
- Conservation of wave power: $P = cgEw$
- $h \downarrow$ then $cg \downarrow$ so $Ew \uparrow$ and $H \uparrow$

Wave crests tend to align with depth contours, because waves travel faster at deeper water

BEACHES

MORPHOLOGY OF BEACHES

Beaches are accumulations of loose sand or pebbles. Much of the energy of waves breaking upon the shore is dissipated in moving sediment, though some is always reflected back out to sea, the proportion depending on the beach slope (Section 1.5.2), and beach profiles can change quite rapidly in response to changes in wave energy

As waves approach the shore, they become unstable and break, forming a breaker zone which can be quite broad, because the waves have a range of wave heights and do not all break at the plunge point (see Section 5.1.1). Landward of the breaker zone is the surf zone, generated by the breaking waves; and finally there is a swash zone, where smaller waves are projected up the beach slope, which is alternately covered by the upsurge of water (the swash) and exposed as the water retreats in the backwash.

In winter more sediment is moved offshore

In general, where there is a seasonal difference between fair weather, swell-dominated waves in summer, and steep storm waves in winter, sediment is moved up the beach slope during summer to build swash bars and/or berms, whereas sediment is removed from the beach during winter and transported seawards, where it may build a longshore bar (Figure 5.1(a)). During the following summer, when swell waves prevail again, sediments are moved back up the beach to reconstruct the swash bars and/or berms. Small, low-amplitude gentle waves and swell thus tend to build up beaches whereas high steep storm waves tend to tear them down

SAND DUNES

Dunes will only develop where winds are favourable and beach sediments fall in the appropriate size range, As with bedload transport by currents (Equation 4.6), the rate of transport of wind-blown sediments is proportional to the cube of the wind speed (but winds need to be much faster than currents to move the sand any distance, because of the lower density of air). Sand dunes tend to migrate inland and can travel long distances if wind conditions allow. They can be thought of as migrating bed forms resulting from wind action (cf. Section 4.4), and they can be stabilized by vegetation, which traps and binds the sand.

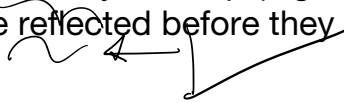
BEACH PROFILES

Where less steep waves break on the beach (collapsing or surging breakers, Figure 1.18), the swash is strong and moves a good deal of sediment up the beach. The backwash is relatively weak because much energy has been used in moving sediment up the beach, and there is more opportunity for water loss by percolation.

Put simply, this means that all waves of a particular height, approaching a beach of given slope, will break at about the same depth, such that for a particular slope, H/d is effectively constant, irrespective of whether they are storm waves or swell. It also means that waves of a given height will break in shallower water (i.e. H/d will be greater) over a gently shelving beach than over a steeply sloping one.

Edge Waves --> As you saw in Figure 1.15, refraction causes oblique waves to become more nearly parallel with the shore as they move into progressively shallower water. If the waves are then reflected, their crests will be rotated away from alignment with the shore as they move into deeper water. With the right combination of water depth and wavelength, the reflected waves can then be refracted back towards the shore, to be reflected again.

Edge waves are best developed where the boundary is steep (e.g. a cliff or harbour wall) and the water is relatively deep, so that the waves are reflected before they break.



SEDIMENT MOVEMENT IN THE BEACH ZONE

Factors that control the rates of movement of sediment onto and off beaches must include the height, speed and steepness of waves, water depth, beach slope, settling velocity of sediment particles, magnitude and direction of local currents, and so on.

This speed is the orbital velocity, and it has a maximum horizontal landwards component immediately beneath the crest of a wave, and a maximum horizontal seawards component just beneath the trough.

such that $\sinh x \sim x$ when x is equal to or less than about 1. For shallow-water waves, $2\pi h/L < 1$ so $\sinh(2\pi h/L) \sim 2\pi h/L$.

The important points to notice in Equation 5.1 are:

- (i) that for waves of a constant period, increased wave height results in increased U_m ; and
- (ii) somewhat counter-intuitively perhaps, that for waves of a given height, U is greater for short-period waves than for long-period waves; i.e. there is an inverse relationship between U_m and T .

Influence of wave characteristics

1 rapidity with which water particles ~accelerate to their maximum horizontal velocity. The acceleration is greater for shorter- than for longer-period waves, and hence so is the frictional (shear) stress at the bed.

2. maximum near-bed horizontal orbital velocities of wave motions are of the same order as the near-bed current speeds required to move sediment particles (e.g. of the order of 0.3 m s^{-1} for 1 mm diameter grains).

3 large storm waves are capable of moving sediment at considerable depths on the continental shelf

4 equation 5.1 shows that there are many combinations of wave period, water depth and wave height which could produce the horizontal orbital velocity necessary to move grains of a given size.

Point -> waves of quite long period can move coarse sediment particles even in relatively deep water (cf. Figure 5.8), so it is hardly surprising that when the waves behave as true shallow-water waves, pebbles and boulders can easily be shifted.

Onshore and offshore movement of sediment by waves

although the amounts of sediment shifted in each direction may be the same, the distribution of particle sizes in each direction is not the same.

On the backwards stroke of the wave, when a water particle is in the trough, it is brought closer to the sea-bed than when it is at the crest of the wave, and frictional retardation at the bed is greater. This means that water particle speeds are not the same in both directions. They are greater during onshore movement, but are maintained for only a short period of time, when coarser sediment is moved shoreward as bedload and finer sediment as suspended load. Speeds are less during offshore movement, so while most of the suspended load and finer bedload are returned seawards, some of the coarser bedload material is left behind. As the slower offshore movements are maintained for longer periods of time (Figure 5.9), there is net offshore movement of fine material, and net onshore movement of coarse material.

High steep waves can transport large pebbles (and even boulders) shorewards, but they will remove smaller gravel particles and sand seawards (again partly in suspension). However, especially during winter storms, the beach tends to be saturated with water, there is less percolation and the backwash is proportionately stronger, so the overall effect is to erode the beach.

Sedimentary features

One of the most common bed form features seen on the beach are wave-generated ripples, which are symmetrical and have long, straight crests that occasionally bifurcate. Wave-generated ripples begin to form as soon as the threshold orbital velocity for grain movement is reached. As orbital velocities increase, the height of the ripples decreases until the sediment moves as a suspended 'sheet flow', backwards and forwards over the sea-bed.

A typically somewhat larger kind of structure commonly seen on beaches are cusps, which are among the most familiar of beach features, being commonest and best developed around high tide at the top of the swash zone. They are series of regularly spaced shallow embayments that are concave seawards and join at sharp seaward pointing 'horns', and are most obvious on shingle beaches.

LONGSHORE SEDIMENT TRANSPORT

The longshore movement of sediments in the beach zone is often referred to as longshore drift, and it can lead to the formation of spits of sand or shingle, elongated in the direction of longshore transport. Spits are generally linear extensions of beaches, sometimes terminating in the open water of bays and estuaries.

COMBINED SEDIMENT TRANSPORT

Most net sediment transport is likely to occur where movement by currents is enhanced by wave motions, because waves are very effective at stirring up sediment on the sea-bed (cf. Figure 4.16). Once waves have lifted sediment into suspension, it can then be transported by currents which would be unable to lift the sediment off the sea-bed by themselves. These currents include both longshore currents and rip currents (see below), as well as tidal currents, and even the slight landwards movement of water caused by the wave drift.

RIP CURRENTS

Rip currents are strong, narrow currents with speeds up to 2 m s^{-1} which flow seawards from the surf zone

The best means of escape is to swim parallel to the shore for a few metres, away from the rip current, before trying to swim shorewards. However, experienced surfers are quite happy to exploit these currents by riding them out to sea.

Rip currents also develop where waves are moving directly onshore. As you know, an increase in wave height along a wave crest occurs when the wave enters shallow water and begins to slow down (Section 1.5). Variations in wave height along a wave crest will occur if one section of the crest encounters shallow water before another.

Just shorewards of the breaking waves, i.e. in the surf zone (Figure 5.1), the average water level falls, and then rises continuously towards the shore, an increase in level known as wave set-up (Figure 5.15(b)) - which can also occur with obliquely breaking waves. Where wave heights are greatest along the wave crest, the wave set-up is also greatest and horizontal pressure gradients are established between regions of higher and lower wave set-up

Increased wave set-up in shallow areas with larger breakers generates long-shore currents towards areas with less set-up, causing seaward jets at their confluence

DYNAMIC EQUILIBRIUM OF SEDIMENT SUPPLY AND REMOVAL

If estimates can be made of the amount of sediment moved by different means in the beach zone, a sediment budget can be drawn up of material imported (e.g. from rivers or cliff erosion) and material exported (e.g. by longshore or offshore transport), and the balance used to determine whether the beach at any one place is undergoing active deposition or erosion

Sediment budgets can also be used to quantify the inputs and outputs for larger-scale sediment transport cells, which are adjacent but discrete segments of coastline where sediment sources and sinks are in overall long-term balance

It can equally well correspond to a situation of net erosion or net deposition - and frequently does.

SUMMARY

1. Beach profiles are controlled by the influence of waves, tidal range, and sediment particle size. The wave zones are the swash zone, the surf zone and the breaker zone. Steep beaches are characterized by berms, shallow ones by swash bars and runnels. Longshore bars may develop seawards of the intertidal zone. Coarse-grained sediments lead to steep beaches because water is readily lost through percolation and the backwash is too weak to move much of the sediment that has been transported up the beach face by the swash. Conversely, fine-grained sediments lead to shallow beaches. Small gentle waves and swell waves tend to build up beaches and steep storm waves tend to tear them down and flatten them.
2. Water particles in shallow-water waves follow orbital paths which become progressively flattened towards the sea-bed. In shallow water, the maximum horizontal orbital velocity and shear stress at the bed increase as the wave height increases and as water depth decreases. The conditions which determine sediment movement for a given grain size may be achieved from many different combinations of wave height, wave period and water depth.
3. Rip currents develop also as a consequence of horizontal pressure gradients between regions having wave set-up of different heights. Convergences of resulting longshore currents lead to the return of water seawards in narrow fast-flowing (rip) currents.
4. The wave power available for longshore sediment transport can be calculated from the wave group speed, average wave height and the angle
5. the wave crest makes with the shoreline. The rate of sediment transport along the shoreline can be estimated using the wave power. Most net sediment transport occurs when movement by currents is enhanced by wave action.
6. Drawing up sediment budgets and identifying coastal sediment transport cells can help to

quantify the dynamic equilibrium of the coastal zone, i.e. whether there is net erosion or net deposition (accretion) of a coastline. Such studies are useful to assess the likely or actual impact of coastal engineering or construction works. Attempts to alter the dynamic equilibrium along one stretch of coastline (e.g. by erosion-prevention measures) are likely to disturb the equilibrium elsewhere, resulting in accelerated (and unwanted) erosion and/or accretion elsewhere. It is probably wiser to let Nature take its course.

What is a beach?

A beach is a landform alongside a body of water which consists of loose particles (Wikipedia)

What defines the landward boundary?

where aeolian (wind/air) processes dominate over water flow processes

→ dunes

What defines the seaward boundary?

sediment not seriously disturbed by wave during fair-weather conditions → water depths of ~10-20 m

VARIATIONS

Build-up by low-energy swell waves (summer); regeneration of berms

- Erosion by steep storm waves (winter); formation of longshore bars

Profiles

Equilibrium profile: balance of onshore sand transport & gravity

- Coarse beaches are generally steeper

SEDIMENT TRANSPORT

Shallow-water environments are the coastal and shallow marine regions which form the interface between the deep ocean basins and exposed land surfaces.

Along passive (aseismic) margins, shelves tend to be broad, whereas along active (seismic) margins they are much narrower, although in the western Pacific some volcanic island arcs enclose quite extensive areas of shelf sea, e.g. the Yellow Sea and East China Sea.

They are generally almost flat, sloping gently seawards from the coast to water depths of between about 100 and 250 m (occasionally as much as 500 m) at the shelf break. The average gradient is only about 0.1°

SEDIMENTS OF SHALLOW-WATER ENVIRONMENTS

Types of weathering

1. Physical (or mechanical) weathering: the fragmentation of rocks, for example by frost-shattering (when water freezes and expands in cracks), or by the cumulative effects of alternate heating and cooling in arid climates at low latitudes, especially in deserts.
2. Chemical weathering: the decomposition of rocks through the reaction of surface waters and groundwater with minerals in rocks, resulting in the formation of new minerals and the removal of material in solution.

In this context, time and distance are important factors: the longer the journey to the sea, the more chance there is for mineral grains to be rounded and reduced in size by abrasion, and sorted according to density and size and shape, and for chemically less stable minerals to break down.

Where there is a plentiful supply of terrigenous sediments, their distribution depends on how they are transported, sorted and distributed by water. Where sediment supply is negligible, however, biogenic sediments (i.e. of biological origin) will accumulate, or older terrigenous sediments (deposited in the past) will predominate.

Tropical latitudes

These are warm, humid regions where chemical weathering would be expected to predominate.

Physical weathering is more characteristic of high latitudes (and high altitudes).

SUPPLY OF SEDIMENT TO THE OCEANS

Figure 3.3 indicates the relative importance of different sediment sources, and shows that rivers supply about 85 per cent of the total solid material entering the oceans.

The dissolved material will be dispersed ocean-wide, but the bulk of the solid products of erosion are deposited in coastal regions and on the continental shelf. At the present day, sediments are supplied to the sea by ice only at high latitudes where glaciers and ice-sheets enter the sea. 95 per cent of such material comes from Antarctica and is deposited within 20 km of the Antarctic coast. By contrast, wind-blown dust is distributed virtually world-wide, though most originates in subtropical latitudes, where Trade Winds blow off continental desert areas such as the Sahara.

Rivers are by far the dominant source of sediments to the oceans, but the supply is by no means evenly distributed around the ocean margins, as Figure 3.4 demonstrates.

Locally, the patterns of sediment supply and distribution can change very rapidly- on a seasonal or even a diurnal time-scale. Storms, floods or droughts can greatly affect the amount of sediments transported to the sea and distributed in the coastal zone, as anyone living near a beach or estuary will know.

Shallow-water environments are transient, existing only for relatively brief periods of geological time.

There is a natural tendency towards equilibrium between the rate at which sediment is supplied to coastal regions, and the redistribution of this sediment by water movement. This equilibrium needs to be understood by marine engineers when they build jetties and breakwaters to protect harbours, construct groynes to prevent beach erosion, or lay submarine cables and pipelines. Otherwise, the equilibrium may be disrupted with disastrous consequences either for the constructions themselves, or for adjacent stretches of coastline, and sometimes even for both.

SUMMARY

1 The sediments of most continental shelves are predominantly terrigenous and consist of rock fragments, quartz sands and clay-rich muds. Where terrigenous sediments are scarce or absent, carbonate sediments of biogenic or inorganic origin may occur, especially in low latitudes.

2 River transport is the most important means of bringing terrigenous sediments to the ocean margins, but ice transport, wind transport and volcanic eruptions may be important locally.

3 Periodic falls in sea-level during the Quaternary resulted in the deposition of river and glacial sediments on areas of the continental shelf which are today covered by the sea. These relict sediments are now being reworked by waves and tidal currents. Changes of sea-level lead to changes in, and migrations of, shallow-water environments on time-scales of hundreds to thousands of years.

PRINCIPLES AND PROCESSES OF SEDIMENT TRANSPORT

Sediment transport and deposition are also more easily studied in shallow water, but the principles governing these processes are as valid in the deep ocean as they are in any estuary or on any beach-or anywhere else where there is moving water.

Plainly, one of the most important parameters controlling the transport and deposition of sediments is grain size, and it is convenient to classify sediment particles on this basis.

Particles too large to be lifted into suspension may be rolled or bounced along the surface by the water flow.

Sliding particles remain in continuous contact with the bed, merely tilting to and fro as they move. Rolling grains also remain in continuous contact with the bed, whereas saltating grains

'jump' along the bed in a series of low trajectories. Sediment particles in these three categories collectively form the bedload. The suspended load consists of particles in suspension, that is, particles that follow long and irregular paths within the water and seldom come in contact with the bed, until they are deposited when the flow slackens. Sliding and rolling are prevalent in slower flows, saltation and suspension in faster flows.

FRICTIONAL FORCES

A boundary layer develops wherever a fluid moves over a surface, whether it be water over the sea-bed, winds over the sea-surface, or syrup over a tabletop.

In theory, provided that no sediment on the bed is moving, the imperceptibly thin layer of water in direct contact with the bed is also stationary - its speed should be zero. However, the layer of water immediately above this does move, albeit very slowly, and slides over the lower layer. With increasing distance from the bed, successive layers of water move a little faster as the effects of friction with the bed decrease. There is thus a velocity gradient- a change of velocity with depth, or velocity shear- and a graph of flow velocity against height above the bed (velocity profile) would resemble Figure 4.2.

To determine whether sea-bed sediment is likely to be moved by a current, we need to know the value of the shear stress actually at the bed. As you will see later, because speed increases systematically with distance from the bed, we can use the rate of increase to determine the shear stress at the bed, rather than measuring it directly (which would be impractical).

The extent to which sediment movement takes place depends upon the degree of turbulence (frictional interaction with the bed is greater when flow is turbulent) and the current shear (i.e. the rate of change of current velocity with depth, Figure 4.2), which determines the shear stress. It also depends upon the nature (composition) and roughness of the bed. Parcels of water follow random eddying paths about the net direction of flow. To deal with turbulence mathematically, the flow velocity can be broken down into three components at right angles to each other (Figure 4.5), each of which at any one time consists of the mean value plus the randomly varying component.

These complications notwithstanding, we can make the general statement that shear stress is proportional to the product of the density of the water and the square of the time-averaged current speed:

From Equation 4.1, if current velocity is doubled, the shear stress will be quadrupled, i.e. it will increase by a factor of four.

The movement of sediment on the sea-bed (Figure 4.1) begins when the shear stress at the bed (τ_0) becomes sufficiently great to overcome the frictional and gravitational forces holding the grains on the bed. This is known as the critical shear stress.

COHESIVE AND NON-COHESIVE SEDIMENTS

In sediments, they tend to form aggregates in which the individual flakes are held together by a combination of electrostatic attraction and the surface tension of the films of water surrounding the flakes. These forces are strong and give muds their glutinous property. Clays increase the overall cohesion of the bed, even when they constitute only a small proportion of the total sediment; cohesion begins to be significant when sediment contains more than about 5-10% of clay by weight. Indeed, the 'stickiness' of many predominantly silty sediments is probably caused by only relatively small amounts of clay. The critical shear stresses required to set cohesive sediments in motion are much greater than might be supposed from their small particle size

They lack the physico-chemical interactions that exist between clay particles, and so are free to move independently. As you might expect, therefore, critical shear stresses required to set non-cohesive sediments in motion decrease with decreasing particle size.

SEDIMENT TRANSPORT, EROSION AND DEPOSITION

1 For particles larger than about 1mm, the 'deposition' curve lies close to but just below the 'erosion' curve. Current speeds required to set these non-cohesive sediments in motion in the bedload are slightly greater than those at which they are deposited, partly because of the inertia of the particles on the bed, and partly because natural sediments are not of uniform grain size, so that larger particles are liable to interlock with and to 'shelter' smaller ones.

2 Particles smaller than about 0.1 mm are typically not transported in the bedload, but taken directly into suspension. When in suspension, however, they can be transported at current speeds very much smaller than those required to erode them, and the difference increases dramatically with decreasing particle size.

EROSION OF COHESIVE SEDIMENTS AND YIELD STRENGTH

The strong binding forces that hold clay particles together mean that cohesive sediments tend to be lifted as clumps or 'flocs', rather than as individual particles, though they generally disaggregate once in suspension. If the muds have become partially consolidated (e.g. on exposed tidal mud-flats, see Chapter 6), erosion occurs following mass failure of the sediment surface, which is tipped off in large lumps - a process that requires very high shear stresses. The cohesion of very fine-grained sediments is influenced by their water content; by the proportion and particle size of clay minerals in them; and by the salinity of both the overlying water and water trapped between the sediment grains

The resistance of mud to erosion can be assessed by its yield strength, which is the maximum shear stress that the sediment can withstand before failure occurs. One method of measuring yield strength is by using a shear vane. Values in excess of 5 N m⁻² may be encountered in the older sediments of estuaries or tidal flats, which tend to be both drier and more compacted than 'fresh' muds.

SHEAR VELOCITY

Molecular viscosity is a constant physical property for any fluid at a given temperature, and in general increases with decreasing temperature, i.e. liquids flow less readily when cooled. The viscosity of water in polar regions is about twice that in equatorial regions, and you may have noticed that water in rivers appears almost 'oily' in cold weather, when it flows more sluggishly.

However, it turns out that at least in the lowermost couple of metres of the boundary layer, flow velocity increases logarithmically with height above the bed (Figure 4.7(a)). Accordingly, if we measure current velocities at various heights above the bed within the lower part of the boundary layer and plot the data on a graph using a log scale for height above the bed, the velocity gradient is a straight line

The relationship between shear velocity and shear stress is given by: $\tau_0 = \rho u_*^2$

The main difficulty with this method is that the theory is strictly valid only for steady flow, and marine currents are rarely steady. We discuss these issues again in Section 4.2.4, but first we consider the intercept of the straight line with the vertical axis at z_0 on the vertical scale in Figure 4.8 (cf. Figures 4.7(b) and 4.9).

THE VISCOUS SUBLAYER

This is the viscous sublayer, no more than a few millimetres thick, which develops only when current velocities are not excessive and where the bed is sufficiently even- i.e. the grain size of sediment

particles and the size of upraised features such as sand ripples are not too great.

As its thickness is inversely proportional to the speed of flow, the viscous sublayer becomes thinner as the current speed increases. This means that a given bottom material may be 'smooth' under a slow current, so that flow is effectively laminar at the bed; but it will become 'rough' under a fast current, so that flow is turbulent right down to the bed.

The near-bed flow is dominated by viscous forces and the flow can be described as hydraulically smooth

Once grain diameters have reached about seven times the notional thickness of the sublayer, they protrude so far into the turbulent layer that the sublayer breaks down altogether and fully turbulent flow conditions extend right down to the bed. Eddies are now able to 'reach' down between the sediment grains and there is much greater potential for sediment movement.

Eventually, with increasing flow speed the shear velocity increases enough to randomly lift grains into suspension and temporarily redeposit them (i.e. the grains move by saltation, Figure 4.1). At still higher shear velocities, the grains are lifted permanently into suspension to be transported as suspended load, provided that eddies have vertical upward velocities exceeding the settling velocities (W_s) of the grains (see Section 4.3.2).

Wherever a sublayer is developed and flow at the bed is smooth, particles can become trapped in the sublayer and be deposited. As mentioned above, sediment particles transported in suspension are kept there only so long as upward velocities in the eddies exceed the settling velocities of the particles. Once the particles get into the sublayer, where the flow is smooth, they can be deposited, even if the mean current speed is well above the 'deposition' curve of Figure 4.6.

VELOCITY PROFILES IN THE SEA

Furthermore, marine currents can have long 'memories', so when a current flowing over the sea-bed encounters a surface with a different roughness, it takes time for the velocity profile in the turbulent boundary layer to adjust. Similarly, changes in current velocity at the surface take some time to be 'transmitted' to the turbulent boundary layer near the bed.

The sea-bed is usually covered with both small-scale and large-scale sedimentary features known as bedforms (see Section 4.4). Probably most familiar are the small-scale sand ripples we have already mentioned, commonly seen on beaches and in estuaries, and formed by waves and currents.

In other words, the ripples provide a partial obstruction to the flow, reducing its capacity to move sediment; this resistance to the flow is sometimes called form drag. In general, the shear stress available to move sediment is less than that calculated from the logarithmic velocity profile, because the transfer of energy and momentum from the moving water results both in movement of sediment and in formation of sediment ripples.

BEHAVIOUR OF NON-COHESIVE SEDIMENTS

In real sediments, the smaller the particles the greater the proportion of clay minerals among them; and we have seen that clay minerals are not rounded but plate-shaped and tend to stick together (they are cohesive materials).

So, for example, from Figure 4.6 we can merely estimate that a current averaging about 0.4 m s^{-1} will set particles averaging 1mm in size in motion as bedload, while average current speeds in excess of 1 m s^{-1} are required to keep such particles in suspension.

RATES OF SEDIMENT TRANSPORT

It is also called the sediment flux. To calculate the total sediment transport rate, we have to consider both the bedload flux and the suspended sediment flux. In general, the amount of suspended

sediment will increase with increasing current speed, as progressively coarser grains are lifted permanently into suspension.

(q_b) is proportional to the cube of the shear velocity, i.e.:

$$q_b \propto U_*^3 \quad (4.6)$$

provided that the shear stress at the bed is greater than the critical shear stress. In addition, since shear velocity is itself related to average current velocity, the rate of bedload transport (the bedload flux) is also proportional to the cube of average current velocity measured at a fixed height above the bed (usually 1m).

Since the energy of the current is proportional to (velocity)², and power is given by (energy) x (velocity), then the power must be proportional to (velocity)² x velocity, i.e. to (velocity)³, as in Equation 4.6. The relationship in Equation 4.6 is important because it means that even very small changes in current speed or bed roughness can have significant effects on the rate of bedload transport. We can illustrate this with a practical example.

One method of measuring bedload transport rates uses the movement of sediment tipples at the sea-bed. These migrate as the sediment moves. So, if we know the size of the tipple, we can calculate the mass of sediment that is moving for a given length of the tipple crest, and then work out the transport rate from how fast the tipple is migrating. To do this, a rod, about one metre long, is held horizontally in the flow about ten centimetres above the bed.

FOR SUSPENDED LOAD

The determination of the rate of suspended load transport is straightforward by comparison with measurement of the rate of bedload transport, but it still has its problems. Current speeds (Figure 4.15(a)) and sediment concentrations (Figure 4.15(b)) are measured throughout the water column and then the suspended sediment flux (through unit area at right angles to the current), q_s , is calculated by multiplying together the two sets of data. Traditional methods for determining the concentration of suspended sediment by direct sampling of the water column are laborious and expensive and often inaccurate.

Yes. Wind-driven currents will tend to flow faster during storms, and where conditions at the sea-surface can affect the benthic boundary layer, sediment transport would be greater at such periods, since doubling the current speed quadruples the shear stress (Equation 4.1), which is the principal factor responsible for sediment erosion.

Thus, in many areas, storms can determine the overall magnitude and direction of sediment transport, even though they may be relatively rare and of short duration.

DEPOSITION OF SEDIMENT

As flow conditions in a shallow marine environment change, it is often useful to know the rate at which sediment is likely to be deposited if current speeds decrease, and the rate at which it will be eroded if speeds increase. The relationship between erosion, transportation and deposition for natural sediments was summarized in Figure 4.6, which takes account of the increasing cohesiveness of sediments as particle size decreases, and of the greater current speeds required to erode such sediments

BED LOAD

Some idea of potential deposition rates from the bedload can be gained if the rate of bedload transport is known. As $q_b \propto u_*^3$ (Equation 4.6), we may assume that the transport rate will decrease (and so the rate of deposition will increase) in proportion to the reduction in u_*^3 and hence also in proportion to the reduction in the cube of the average actual current velocity $E^{1/3}$.

SUSPENDED LOAD

Moreover, as a current slows down, suspended particles of a given grain size do not all reach the bed at the same time because they will be distributed at different depths in the water column (Figure 4.15(b)); hence the rate at which suspended sediment is deposited depends on more than just the decrease in current speed. The time that particles take to settle will depend to a large extent on their settling velocities, W_s (Figure 4.6(c)), and on the degree of turbulence in the water column- and while the particles are settling, they continue to be transported in the direction of net current flow. Since very small particles settle significantly more slowly than large ones (especially if they are flaky), they will eventually reach the bed some distance from where they began to settle, i.e. there is a settling lag.

In short, the smaller the particle size, the slower the current can flow while continuing to transport sediment, the greater the settling lag, and the slower the rate of deposition. As you will see in Chapter 6, for muds to be deposited at all, either the water must be completely still for long periods, or the particles must somehow be brought together into larger aggregates with greater settling velocities. Other factors affecting rates of deposition of suspended sediment include: the degree of bed roughness (which generates near-bed turbulence); the extent of resuspension of sediment by marine organisms; and occasional brief episodes of erosion during an overall period of net deposition.

BED FORMS

The sea-bed is rarely flat: it is usually covered with both small- and large- scale sediment features. The most familiar of these are probably the small- scale sediment ripples formed by waves and currents, commonly seen on sandy beaches and in the silts and muds of estuaries and tidal flats.

The oscillating water flows in wave motions make sediment ripples with generally symmetrical cross-sections, but current-produced bed forms are usually asymmetrical with the steeper slope facing downstream.

- Current ripples commonly develop where relatively slow currents move over sediments finer than about 0.6 mm grain size, and water flow is only slightly disturbed above the ripples (Figure 4.18(a)). It seems they can form only where the viscous sublayer is not completely broken down by turbulence and grain roughness (Figure 4.10). Since the thickness of the sublayer decreases as the speed of a current increases, it follows that ripples are unlikely to develop at fast current speeds or where there is coarse sediment on the bed.
- At higher current speeds, and in coarser sediments, somewhat larger bed forms known as megaripples are produced (Figure 4.19(b)). These are up to a metre or more in height, and often have wavelengths of several metres or even tens of metres.
- At current speeds of up to about 0.75 m s⁻¹, sand-waves can occur extensively. For example, they cover an area of 15 000 km² off the Netherlands coast. Sand- waves may reach 18 m high and have wavelengths of nearly a kilometre, though most are much smaller (Figure 4.20, overleaf). Individual sand-waves may migrate in the direction of the mean or residual current, at rates of between 10 and 150 m per year.
- All of these bed forms are transverse to the current flow and develop where surface current flows do not exceed about 0.5-0.75 m s⁻¹. At greater current speeds, bed forms become linear and approximately parallel to the flow direction. Where there is abundant sand on the sea-bed and mean tidal current flows are more than about 0.5 m s⁻¹, sand-banks can be expected to develop. These large submarine features are found on many continental shelves, particularly in shallow water, and may reach 80 km in length, 3 km in width and tens of metres in height, oriented parallel to the tidal currents or converging in the down-current direction
- The furrows are probably passageways for removal of sand, as linear sand ribbons often extend far down-current from them, reaching 15 km in length and 200 m in width, but only about 1 m in

thickness. Where current speeds exceed about 1.5 m s^{-1} , scour hollows may develop.

SUMMARY OF CHAPTER

1. Friction between flowing water and the sea-bed generates a boundary layer in which turbulent flow dominates, except very close to the bed. Movement of sediment (erosion) occurs when the shear stress generated by the frictional force of water flowing over the sediment overcomes the force of gravity acting on the sediment grains and the friction between the grains and the underlying bed. Shear stress is proportional to the square of the mean current speed (and to the density of the water). Movement of grains of a given size begins when the shear stress at the bed reaches a critical value (critical shear stress). Particles larger than about $0.1\text{--}0.2 \text{ mm}$ in diameter are initially moved as bedload, then lifted intermittently into suspension, and as current speeds increase further they are finally lifted permanently into suspension. Grains smaller than about 0.1 mm are lifted directly into suspension as soon as the critical shear stress is reached.
2. Cohesive sediments contain a high proportion of fine-grained clay minerals and are more difficult to erode than non-cohesive sediments, which consist mostly of quartz grains. For cohesive sediments, the smaller the particle size, the greater the current speed required to erode them.
3. When current speed is plotted against the height above the sea-bed (as the vertical axis) on a log-linear graph, the inverse velocity gradient $d \log z / dE$ is linear. The slope of the line is used to calculate the shear velocity, and the intercept of the line with the depth axis gives a measure of the bed roughness length (z_0) which increases as the sediment grain size increases;
4. When water flows over a smooth (very fine-grained) bed, the lowermost water layer appears to flow in a laminar fashion, forming a viscous sublayer only a few millimetres thick, which decreases in thickness as flow speed increases (and as viscosity decreases with increasing temperature). Suspended particles which settle into the sublayer are subjected only to laminar flow and are soon deposited. When water flows over a coarse-grained bed, or at high current speeds, grains protrude through the viscous sublayer, breaking it down, and turbulent flow extends to the bed, giving rise to greater potential for sediment movement.
5. The rate of bedload transport is proportional to the cube of the shear velocity ($q_b \propto u_*^3$) and hence also to the cube of the average current speed, but it is difficult to measure directly in the sea. The rate of suspended load transport can be calculated from the product of the current velocity and the sediment concentration.
6. Deposition of the bedload begins when the current speed falls so that the shear stress at the bed is only a little below the critical shear stress required to start the sediment moving. The rate of deposition of the bedload is proportional to the reduction in the cube of the average current velocity, q_a^3 . The rate of deposition of the suspended load varies according to sediment grain size. The settling lag of fine suspended particles means that they may reach the bed well after they began to settle from suspension. The rate of deposition of grains of a given size in the suspended sediment load depends both on the vertical distribution of sediment concentration above the bed, and on the settling velocity of the grains.
7. Bed forms
Produced by waves are symmetrical, those formed by currents are asymmetrical. The type of bed form depends mainly on current speed and sediment grain size and partly also on water depth. Small-scale current ripples form at relatively slow current speeds and where sediment is finer than about 0.6 mm grain diameter. As current speeds increase or where sediments are coarser-grained, larger-scale megaripples are formed; and sand-waves develop where sand is abundant. At current speeds of more than about 1 m s^{-1} , bed forms such as sand-banks and sand ribbons develop parallel, rather than transverse to, the current flow. At still higher current speeds, erosional features such as furrows and scour hollows develop.

